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THE HEIGHT OF TROPICAL CYCLONES AND OF THE "EYE" OF THE STORM

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The opinion has often been expressed that tropical cyclones are very flat disturbances. In Hann's "Lehrbuch der Meteorologie" (4th edition, 1926), page 620 fig., it is stated that they are mostly phenomena restricted to the lowest layers of the atmosphere, or at least their active vortex does not reach very high; similarly, Mrs. E. V. Newnham¹ assumes that the height of tropical cyclones is small. This opinion is based on the observation that tropical cyclones decrease rapidly in intensity when passing over land, particularly in crossing even small mountain ranges. Such a case was reported by John Eliot² for the Backergunge cyclone of October 1876, which was "completely broken up by the hills (3,000–6,000 feet) in eastern Bengal and South Assam." Other observations are cited by Mrs. Newnham.

The reported rapid disintegration of the Backergunge cyclone over hills apparently does not hold for typhoons in general, however. Unfortunately there is no observational material for this cyclone which could enable us to study this phenomenon more closely. Cline³ has shown that in the tropical cyclone of August 17–22, 1915, the amount of rainfall decreased by about one-half when the cyclone reached the low mountainous region of western Arkansas (1,500–2,500 feet) but markedly increased again upon reaching the level lands of the central Mississippi Valley. Likewise the observations mentioned by Mrs. Newnham seem to show that cyclones regenerate after having crossed mountain ranges. We may, therefore, conclude that the influence of mountain ranges of moderate height, while diminishing the intensity of the cyclone to a certain degree, does not dissipate the whole disturbance. It seems rather that the cyclone regains its former intensity after having crossed the obstacle. Sir Napier Shaw⁴ has expressed the opinion that a mountain range might possibly cut off the lower part of a cyclone, so that this part would have to be rebuilt after the cyclone had reached level country again. Thus the annihilation of tropical cyclones by mountain ranges does not really give an indication of their height; and the not infrequently reported regeneration of cyclones after they have crossed mountain ranges would rather indicate that the dynamical conditions extend to high altitudes. Without going into more detail it may be said that the common arguments in favor of a small vertical extension of tropical cyclones seem rather weak. On the other hand, good reasons can be brought forth in favor of the view that tropical cyclones have about the same height as those of middle latitudes.

¹ Newnham, Mrs. E. V.: Hurricanes and Tropical Revolving Storms (with an introduction on the Birth and Death of Cyclones, by Sir Napier Shaw). Geophys. Memoir No. 19, Meteorological Office, London, 1923, p. 232.

² Eliot, John: Handbook of Cyclonic Storms in the Bay of Bengal. Calcutta, 1900, p. 228.

³ Cline, I. M.: Tropical Cyclones. New York, 1926, p. 100.

HEIGHT OF PRESSURE EQUALIZATION AND REQUIRED LAPSE RATES

W. Köppen⁴ has pointed out that an unreasonably high temperature would be required in the central part of a hurricane if the pressure difference between center and outer parts were to be smoothed out at an altitude of, say, 3 km. It should be mentioned in this connection that the elevation where the pressure is uniform, or, more exactly, equal to the general pressure distribution in the region, cannot really be regarded as the upper limit of the cyclone. Above this altitude the pressure gradient may, and probably does, have the opposite direction, in order to allow the air which flows inward in the lower strata of the hurricane to flow out again. However, we may take the height of the surface of pressure equalization as an index of the vertical extent of a hurricane.

We shall now try to obtain some information about the height of a hurricane by computing the change in the vertical lapse rate of temperature necessary to smooth out at different elevations the decrease of pressure observed at the ground. For simplicity we shall use for this computation the mean temperature of the whole air column, as the results thus obtained differ but little from those which the exact barometric formula for a linear lapse rate would give.

Let P be the pressure at the ground; p , the pressure at the level, h , of uniform pressure; T_m , the mean temperature; g , gravity; and R , the gas constant for air ($g/R = 3.042 \times 10^{-2} \text{ } ^\circ/m$). Symbols which refer to values in the undisturbed area outside the cyclone will be indicated by the index 1, while the index 2 refers to the interior of the cyclone. Then we get as the condition that the pressure at the altitude h be uniform,

$$p_1 = p_2,$$

or by means of the barometric formula,

$$P_1 e^{\frac{-gh}{RT_1}} = P_2 e^{\frac{-gh}{RT_2}},$$

which gives for the difference of the mean temperatures⁵

$$\frac{T_2 - T_1}{T_1 T_2} = \frac{R}{g} \frac{1}{h} \ln \frac{P_1}{P_2}, \quad (1)$$

or

$$T_2 = \frac{T_1}{1 - \frac{R}{g} \frac{T_1}{h} \ln \frac{P_1}{P_2}}$$

⁴ Köppen, W.: Met. Zeit., 1920, vol. 37, p. 328.

⁵ It would be shorter to derive an equation for the difference of the mean temperatures by differentiating the barometric formula. However, the actual differences under examination are too large to allow the application of such a formula.

If we know the mean temperature of the air column, its lapse rate can be assumed approximately as

$$\alpha = 2 \frac{T_o - T_m}{h} \quad (2)$$

Before we give a numerical discussion of the lapse rates necessary to compensate the surface pressure differences at different altitudes, the behavior of the surface temperature in different parts of the center of a tropical cyclone should be discussed. The temperature records of some cyclones traveling over the southern United States, published by I. M. Cline,⁶ show no evidence of a higher surface temperature toward the center. In some instances the thermograph shows a slightly higher temperature while the station is near the center of the storm; in other cases, however, the temperature is lower. It is true that these cyclones, having traveled over land for a longer or shorter time, have partly lost the characteristic qualities of tropical hurricanes; but since we have no other observations we shall accept Cline's opinion that these temperature changes result mainly from the cloudiness which cuts off the insolation. We therefore assume first that the surface temperature is uniform, and choose the following numerical values: The surface temperature may be $300^\circ A$ ($81^\circ F.$), and the lapse rate outside the cyclonic area $\alpha_1 = 0.6^\circ/100m$; that would make $T_{m1} = 291^\circ$ for an air column 3 km high, $T_{m1} = 282^\circ$ for an air column 6 km high, and $T_{m1} = 270^\circ$ for an air column 10 km high. The surface pressure outside the cyclonic area may be 1,010 mb (29.82 in.); the pressure in the center may be 970 mb (28.64 in.). This value for the lowest pressure is certainly a very conservative assumption; however, this will only strengthen our argument. The formulae (1) and (2) show that it would require a mean temperature of $T_{m2} = 328^\circ$ to smooth out the pressure drop at the surface at an elevation of 3 km. That would mean an increase in temperature with elevation, of $1.87^\circ C./100 m$ in the central parts of the cyclone. We have no means of accounting for such an exceptionally high temperature in hurricanes, and therefore the conclusion is unavoidable that the surface of pressure equalization must be much higher. If it should be at an altitude of 6 km, the mean temperature in the central part should be $T_{m2} = 299^\circ$, which corresponds to a vertical temperature decrease of $.03^\circ C./100 m$, or an almost isothermal state. If we put $h = 10 km$, we get $T_{m2} = 279^\circ$, and $\alpha_2 = .42^\circ C.$ per 100 m. It is highly improbable that there is an isothermal condition up to 6 km in any tropical cyclone and it therefore seems reasonable to assume that the surface of pressure equalization is at an elevation of 10 km rather than 6 km.

This opinion is strengthened by the results of Horiguti's⁸ investigations of the typhoons of the Far East. In these cyclones the temperature decreases slightly toward the center on land; the difference between the outer and inner parts is about $2^\circ C.$ At sea the temperature increases slightly toward the center by about the same amount, but only on the front side, remaining fairly constant over the entire rear side. We should also keep in mind here that these typhoons after traveling over land have lost many of their characteristic qualities.

OBSERVED LAPSE RATES IN TYPHOONS

Especially interesting in Horiguti's investigations of the temperature distribution in typhoons is the fact that he could also obtain some determinations of vertical

temperature gradients from observations on Mount Tukuba (869 m.), Mount Bessi (788 m.), and Mount Ibuki (1,376 m.). While it is true that temperature gradients obtained from mountain observations are rather unreliable indicators of the temperature gradients in the free atmosphere, since they are very much influenced by the topography, we shall discuss them here for lack of other observations. Horiguti summarizes these observations by stating that the lapse rate is about 0.6° or $0.7^\circ/100 m$ or even greater in the outer part of a typhoon, and about $0.5^\circ/100 m$ in the inner part. If we may assume that these mountain values for the lowest layers are representative of the values of the lapse rate in the free atmosphere, we see that they indicate a higher mean temperature in the central parts of the hurricane. However, this higher mean temperature is not nearly sufficient to cause a uniform pressure distribution in the lowest layers of the atmosphere, as we shall presently show, for even the most favorable of Horiguti's cases. Horiguti's material contains only two cases in which lapse rates could be determined when the pressure values were less than 740 mm (29.13 in.). In one of these cases, however, the lapse rate was practically the same both inside and outside the central part of the typhoon, so that only at greater elevations over the mountain can the temperature compensation of the drop in surface pressure take place. The other observation was made during the typhoon of August 4, 1920, at Mount Ibuki and at Hikone (88 m, 23 km to SW.). The following values were obtained:⁹

TABLE 1.—Lapse rates in a typhoon

Date	Hikone		Mount Ibuki		Lapse rate	Distance from center
	P	T	P	T		
1920	mm	° C.	mm	° C.	° C./100 m	km
Aug. 1.....	750.1	26.8	647.1	17.4	.73	1,000
Aug. 2.....	48.6	25.9	45.3	17.7	.64	800
Aug. 3.....	44.8	26.1	41.8	17.6	.66	700
Aug. 4.....	35.8	22.9	33.9	15.7	.48	200
Aug. 5.....	38.2	22.1	35.4	13.4	.68	400
Aug. 6.....	42.7	22.9	40.0	14.3	.67	(¹)

¹ Outside of the depression.

If we choose, for instance, the values at 800 and 200 km from the center, we can reverse our previous question and compute the elevation at which the pressure will be uniform under the assumption that the lapse rates found between mountains and valleys hold throughout the whole layer under consideration.

By means of formulae (1) and (2), we obtain the equation

$$\left[\frac{g}{R} \frac{\alpha_2 - \alpha_1}{2} + \frac{\alpha_1 \alpha_2}{4} \ln \frac{P_1}{P_2} \right] h^2 + \left[\frac{g}{R} (T_1 - T_2) - \frac{\alpha_1 T_2 + \alpha_2 T_1}{2} \ln \frac{P_1}{P_2} \right] h + T_1 T_2 \ln \frac{P_1}{P_2} = 0 \quad (3)$$

where the indices 1 and 2 refer to the values in the inner and outer parts of the cyclone respectively. With the values published by Horiguti, equation (3) gives for h , 8,950 m. If instead of the values for 800 km, we take these for 700 km distance from the center, we find $h = 7,520 m$. Both values indicate as clearly as the previous example that the drop in pressure cannot be smoothed out in the lowest layers of the atmosphere, even though we could use only data from the less violent parts of the typhoon. Here again we have used temperature distribution data which were obtained in hurricanes

⁶ I. M. Cline, loc. cit., pp. 205, 206, New York, 1926.

⁸ Yosiki Horiguti, On the Typhoons of the Far East, Part II. Mem. Imp. Mar. Obs. Kobe, Japan, 1927. Vol. III, no. 2.

⁹ Horiguti, loc. cit., p. 61.

weakened by moving inland, but it can be seen from the foregoing discussion that this unreliability of the data does not affect our argument in favor of a vertical extent of the hurricane through at least the whole troposphere.

WARMTH AND DRYNESS OF THE "EYE" OF THE MANILA TYPHOON

So far we have assumed that the surface temperature is uniform throughout the area of a cyclone, an assumption which was based upon I. M. Cline's work and upon Horiguti's observations. There are cases reported, however, when the temperature rose considerably with the approach of the storm center. In the Manila typhoon of October 19-20, 1882,¹⁰ the temperature was fairly uniform at about 24° C. until the center approached. In the calm the temperature rose suddenly to 31.6°; then fell to about 25° after the center had passed. If we assume this rise in temperature of 8° is characteristic of the difference in the mean temperature of the air in the outer and inner zones of the hurricane, we can see, without computation, from our first example that the surface of equal pressure can hardly be below 10 km. We assumed there a drop in pressure of 40 mb, which corresponds to the values in the Manila typhoon, and we saw that an increase of 9° in the mean temperature is required to smooth out this effect at 10 km elevation. The relative humidity, which was almost 100 percent immediately before and after the calm, dropped to 49.7 percent.¹¹ The rise in temperature may be due to insolation, according to Algué, in spite of a continuous "veil of condensed vapor." If the change in relative humidity had been caused by insolation of the same air mass that was present before the calm, we should expect to observe a drop in relative humidity to about 61 percent; but the air in the center is very much drier, indicating that in the eye of the storm we have a downward current which brings warm and dry air to the ground. This air in the center probably comes from the surrounding regions of the cyclone, and has lost part of its moisture content by precipitation during previous ascent.

The innermost part of the cyclone, the calm center, seems to be composed of warmer and drier air than the outer part. If we assume that the rise in temperature observed at the ground is characteristic of the difference between the mean temperatures of the air in the calm and in the outer part, the height of the surface of pressure equalization will be about 10 km.

NEGLIGIBLE EFFECT OF MOISTURE AND VERTICAL ACCELERATIONS ON PRESSURE DISTRIBUTION

A few words must be added about the influence of the moisture content and the effect of vertical accelerations on the pressure distribution. It is generally recognized that both factors are very small, but they will be discussed here for the sake of completeness. It is known that moist air under the same pressure and temperature is lighter than dry air and that it is therefore possible in many problems of atmospheric statics to treat the actual moist air as dry air of a higher, "virtual" temperature T' . If T is the actual temperature, e the vapor pressure, p the air pressure, we have

$$T' = T \left(1 + 0.377 \frac{e}{p} \right)$$

A temperature of 27° C. and saturation, for instance, would give $e=35.6$ mb and with $p=1010$ mb, $T'=300^\circ +$

4°. This value of 4° C. is fairly high and will seldom occur actually. However, let us assume, for the sake of the argument, that the moisture content is so high that the mean temperature of the whole air column is increased by the same amount. (It is hardly necessary to mention that this assumed condition is impossible, since the specific humidity decreases with elevation.) If we return to our first example, we see that we could reduce the actual mean temperature in the inner part by 4°, provided that the outer regions have no moisture content, while the interior part contains the excessive amount just assumed; but even then we should find a lapse rate of only about 0.17°/100m, if we assume that the horizontal pressure differences are smoothed out at 6 km elevation. Furthermore, the existence in the strongest cyclones of an "eye" with its dry air shows that here the influence of the moisture content in the central part can hardly contribute to the decrease of the pressure.

Still more negligible is the influence of vertical accelerations. Let ρ be the density, p the pressure, z the height, g the acceleration due to gravity, and w' the acceleration of the air (which we assume to be constant). We see, then, from the equation for the vertical motion,

$$-\frac{1}{\rho} \frac{\partial p}{\partial z} = g + w' = \bar{g}$$

that the effect of the vertical acceleration can be interpreted as a slightly changed gravity acceleration, somewhat larger with upward acceleration and smaller with downward acceleration. Therefore the influence of vertical acceleration upon the surface pressure, under the assumption that the pressure p at an elevation z remains unchanged, is easily obtained by differentiating the barometric formula with respect to g . If we again write w' for dg we obtain

$$\frac{dP}{P} = \frac{z}{RT_m} w'.$$

In order to judge of the possibility of such values of w' as we might compute by assuming certain values for dP , P , z and T_m we must remember that

$$w = \sqrt{2w'z} = \sqrt{2RT_m \frac{dP}{P}}$$

provided that w' is constant. If we put $dP=1$ mb, $P=1010$ mb, $T_m=267^\circ$ we obtain for different altitudes:

TABLE 2.—Vertical acceleration and velocity required for an appreciable effect on pressure

z	3.10 ³ m	6.10 ³ m	10.10 ³ m
w'	2.5.10 ⁻²	1.26.10 ⁻²	.76.10 ⁻² m/sec. ²
w	12.3	12.3	12.3 m/sec. ²

These values are by far too high, so that evidently the vertical acceleration has practically no influence on the pressure distribution even in tropical cyclones. It might be added that in tornadoes, owing to their much more violent vertical motions, the influence of vertical motions on the pressure distribution is probably appreciable; but this vertical acceleration could not take care of even part of the drop in the surface pressure toward the center. Indeed, since we have an increased upward motion and therefore an increased upward acceleration toward the center, we should expect a rise in pressure in this direction if everything but the vertical acceleration remains unchanged.

¹⁰ J. Algué, S. J.: The Cyclones of the Far East, 2d ed., Manila, 1904.

¹¹ Or 43 percent, according to the observations of Fr. Faura, cit. by Algué, loc. cit., p. 57.

THE FUNNEL OF WARM DRY AIR IN A HURRICANE

We have heretofore been concerned only with the pressure differences between two particular points, namely, the center and a place at the outer limit of the hurricane. Going one step further, if we take for granted that the central air column of the hurricane consists of a drier and (which is more important at present) warmer air mass than the outer part, we may assume that the boundary between these two air masses has a form such as indicated in fig. 1, a funnel, the surface of which curves outward

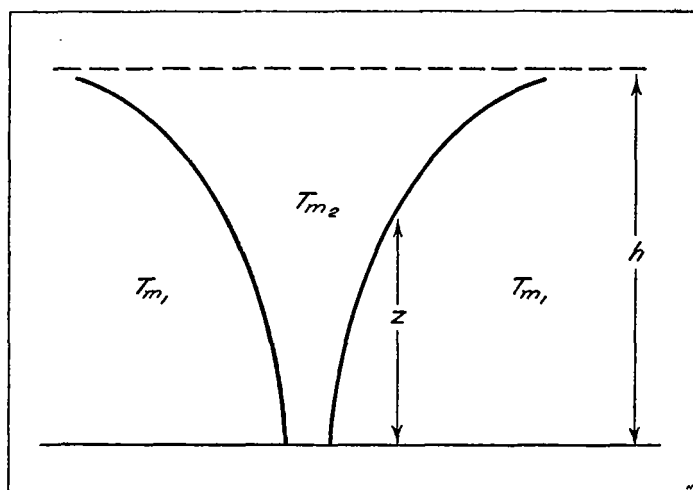


FIGURE 1.—Scheme of the center of a tropical cyclone.

with elevation. Such a shape has been suggested by other writers. We can now try to get more information about the shape of this funnel by assuming two different mean temperatures, T_{m2} for the air inside the funnel, and T_{m1} in the outer part. Then we can compute at what altitude the colder air of temperature T_{m1} has to be replaced by the warmer air of temperature T_{m2} in order that at an elevation h the pressure will be uniform. Since the pressure at the elevation z is

$$p_z = P e^{-\frac{g}{R} \frac{z}{T_{m1}}}$$

and the pressure at the elevation h is

$$p_h = p_z e^{-\frac{g}{R} \frac{h-z}{T_{m2}}} = P e^{-\frac{g}{R} \frac{z}{T_{m1}} - \frac{g}{R} \frac{h-z}{T_{m2}}}$$

we get

$$z = \frac{R}{g} \frac{T_{m2} T_{m1}}{T_{m2} - T_{m1}} \left(\ln \frac{P}{p_h} - \frac{gh}{RT_{m2}} \right) \quad (4)$$

For a numerical example, let

$$T_{m1} = 267^\circ, T_{m2} = 275^\circ, P_1 = 1,010 \text{ mb}, P_2 = 970 \text{ mb}$$

These figures are chosen so as to agree with the data for the Manila typhoon of October 20, 1882. The mean temperature, T_{m1} , is computed on the assumption of a surface temperature of 24°C ., a lapse rate of $0.55^\circ \text{C}/100 \text{ m}$ and a height h of 10,860 m for the surface of pressure equalization. This height can at once be found by means of formula (1).

A second possibility for the determination of h is given by formula (4): The elevation z of the boundary must become 0 at the beginning and end of the calm zone. We can therefore put $z=0$ in formula (4) and introduce for P the pressure value observed at the borders of the calm. We thus arrive at two equations for the three quantities h , T_{m1} , T_{m2} . If we choose T_{m1} , for instance, according to

the surface observations and with reasonable assumptions as to the lapse rate in the outer part we can compute h , the surface of pressure equalization, as we have just done for the Manila cyclone, and in addition T_{m2} , the mean temperature in the central part. In practice, however, a deficiency of our theory manifests itself here. According to Algué the pressure was not constant during the passage of the calm, but had already begun to rise when Manila entered the calm zone, showing that the center of motion lags behind the center of pressure. That may be due to the progressive motion of the cyclone as a whole,^{12 13} which we disregarded in deriving our formulae in order to simplify matters and to work out the essential points more clearly. We shall therefore have to be content to assume plausible values for T_{m1} and T_{m2} and compute h only. The results which we obtain in this way show that we are not very far from the truth in spite of our neglect of the progressive motion of the cyclone. It will, however, be worth while to keep in mind that with a more detailed theory we would be able to restrict our assumptions further.

Having found h we can at once determine P_h , the pressure at the surface of pressure equalization. Then we can find from formula (4) the elevation z of the boundary between the inner and outer air masses as a function of the surface pressure P . The result is partly given in the following table, and is plotted in figure 2. As abscissa we have chosen the distance from the center instead of the surface pressure. This could easily be computed, since Algué gives the velocity of the hurricane at 19 nautical miles (35.2 km per hour) and states that the center passed Manila on October 20, 1882, at 11:52 a. m. This speed could have not been quite constant, hence our abscissa may not be exact; however, we can be sure that its accuracy is within reason.

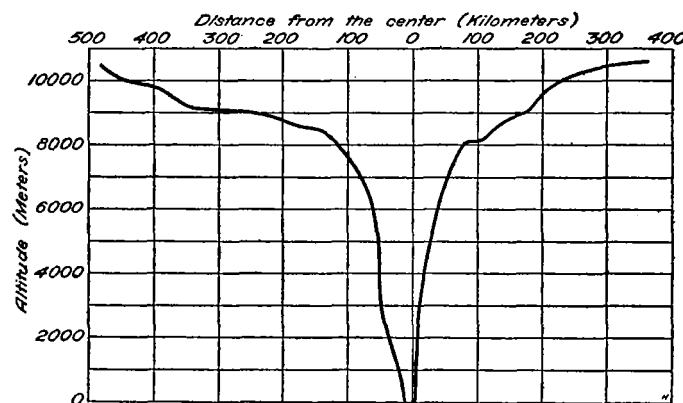


FIGURE 2.—Shape of the funnel of the Manila cyclone, October 19-20, 1882.

Table 3 and figure 2 give the funnel-shaped form of the "eye" of the storm. The two sides of the boundary have not exactly the same shape. The boundary is steeper on the rear side than on the front side of the cyclone. That is, of course, closely connected with the fact that the isobars, as happens frequently, are crowded together on one side of the cyclone; but this cannot be regarded as an explanation; it is just another manifestation of the same phenomenon, since we used the surface pressure for the calculation of the height of the boundary. The crowding of the isobars in one direction is probably in many cases an effect of the progressive motion of the cyclone as a whole.

¹² In other cases, however, the center of pressure lags behind the center of motion. It is therefore possible that our simple kinematic explanation does not hold even in the case of the Manila cyclone, but that other influences are at work which bring about the displacement of both centers with respect to each other.
¹³ cf. Th. Hesselberg, Über das Verhältnis zwischen Druckkraft und Wind. Geofys. Publik., Vol. IX, No. 8, Oslo, 1932.

TABLE 3.—Height of the boundary of the "eye" of the Manila cyclone, Oct. 19–20, 1882

Before passage of center		After passage of center	
Distance from center	Height of boundary	Distance from center	Height of boundary
km.	m.	km.	m.
312	9,100	40	5,830
243	9,000	75	8,050
171	8,530	110	8,150
101	7,590	180	9,060
66	6,410	251	10,100
48	3,460	321	10,500
30	1,740		

We do not claim that the boundary of the calm zone as represented by figure 2 is in all respects a true boundary in the sense that no air can cross this surface. It might well be that only the inner steep part is a boundary in this hydrodynamical sense of the word, while the more horizontal part is rather a transitional zone where the air changes its qualities more rapidly.

Formula (4) may also be derived by means of the relation between the inclination of a surface of discontinuity and the velocity and density distribution on both sides. This deduction shows that formula (4) is also in agreement with reasonable assumptions about the wind distribution in a tropical hurricane.

SOUNDING-BALLOON OBSERVATIONS AT OMAHA, NEBR., DURING THE INTERNATIONAL MONTH, JANUARY 1934

By J. C. BALLARD

[Weather Bureau, Washington, D. C., January 1935]

The Weather Bureau conducted a series of sounding-balloon observations at Omaha, Nebr., during the international month, January 1934. The observations were made in cooperation with the International Commission for the Exploration of the Upper Atmosphere, and the program followed was similar to that for other series of this nature except for the addition of some special observations. The latter included a total of 5 observations on the 21st and 22d, 2 on the 24th, and 5 on the 30th.

It was intended to make the special observations during the passage of a low-pressure area over the station, to obtain a cross-sectional picture of the meteorological conditions. No satisfactory condition occurred during the month, however, so the instruments for special use were released as indicated below. Those of the first set (on the 21st and 22d) were released in the southern part of a low, those of the 24th were made during the advance of a cold wave, and those of the 30th were made to determine whether a slower rate of ascent would result in greater altitudes being reached by the balloons before bursting. The reason for expecting better altitude performance from a nearly floating balloon lies in the fact that at low temperatures more time is required for the rubber of the balloons to stretch to the fullest extent than at higher temperatures. The best altitude performance from a slowly rising balloon probably would be obtained in the daytime, but the temperature record would then be of little value.

For a description of the instrument used, the method of attachment, etc., and a record of previous sounding-balloon series in this country, the reader is referred to the following issues of the MONTHLY WEATHER REVIEW, in which either references or descriptions will be found: January 1932; February 1934; April 1934.

Most of the observations were made during daylight hours, so that the balloons could be followed by theodolite; and even though they were made as late in the day as practicable, the temperature records probably were affected by insolation. However, the amount of this effect could not be determined because of lack of a sufficient number of night flights (1).

Detailed data obtained from the observations will be published by the International Commission for the Exploration of the Upper Air.

A summary of the individual observations will be found in table 1. Altogether 46 instruments were released and all but two were found. Two of the records were de-

stroyed by the finders. Of the remaining 42 records, 38 were good; and at least 1 good record was obtained on all but 3 days during the month.

The highest altitude reached was 29.3 km on the 10th, and a height of 20 km was exceeded seven times. The average height reached in those flights in which the record was good up to the maximum altitude was 17.5 km.

The greatest height of the tropopause observed during the month was 14.2 km, on the 10th; and the lowest, 7.9 km, on the 12th. The average height was 11.1 km.

Figure 1 shows the average temperature curve for the month, obtained by the difference method from the regular flights made near 4 p. m. (the special observations were not considered) and the mean temperatures for the month obtained from the airplane observations made near 4 a. m. It will be noted that the morning temperatures observed by airplane averaged higher above 800 meters (m. s. l.) and lower above 3,000 meters (m. s. l.) than the afternoon temperatures observed by sounding balloon, in spite of the fact that due to diurnal variation the temperature below 1,500 meters

(m. s. l.) averages lower in the morning than in the afternoon (2).

As a further check on the agreement between the temperatures observed by the two methods, the Polar Year observations were used (1). On 2 days per month during the Polar Year, sounding-balloon observations were made near 6 a. m. On each day on which one of these sounding-balloon observations was available, to-

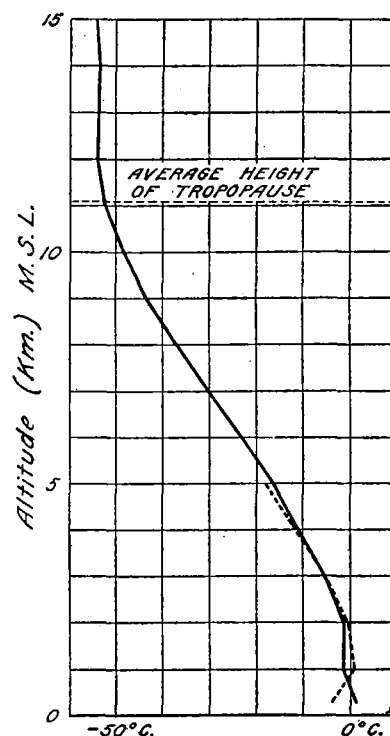


FIGURE 1.—Mean temperature curve for January 1934 at Omaha, Nebr. Solid line based on late afternoon sounding-balloon observations and dashed line on early morning airplane observations.